Impact of Eurasian Spring Snow Decrement on East Asian Summer Precipitation

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ABSTRACT

In this study, the relationship between Eurasian spring snow decrement (SSD) and East Asian summer precipitation and related mechanisms were investigated using observational data and the Community Atmospheric Model, version 3.1 (CAM3.1). The results show that a west–east dipole pattern in Eurasian SSD anomalies, with a negative center located in the region between eastern Europe and the West Siberia Plain (EEWSP) and a positive center located around Baikal Lake (BL), is significantly associated with East Asian summer precipitation via triggering an anomalous midlatitude Eurasian wave train. Reduced SSD over EEWSP corresponds to anomalously dry local soil conditions from spring to the following summer, thereby increasing surface heat flux and near-surface temperatures. Similarly, the increase in SSD over BL is accompanied by anomalously low near-surface temperatures. The near-surface thermal anomalies cause an anomalous meridional temperature gradient, which intensifies the lower-level baroclinicity and causes an acceleration of the subtropical westerly jet stream, leading to an enhanced and maintained Eurasian wave train. Additionally, the atmospheric response to changed surface thermal conditions tends to simultaneously increase the local 1000–500-hPa thickness, which further enhances the Eurasian wave train. Consequently, significant wave activity flux anomalies spread from eastern Europe eastward to East Asia and significantly influence the summer precipitation over China, with more rainfall over northeastern China and the Yellow River valley and less rainfall over Inner Mongolia and southern China.

1. Introduction

Variations in Eurasian snow cover in winter and spring play a major role in the East Asian climate and precipitation during the following seasons (Gong et al. 2003, 2004; Wu and Kirtman 2007; Zhang et al. 2008; Wu et al. 2009a; Zuo et al. 2011, 2012; Mu and Zhou 2012, 2015; Zhang et al. 2013; and many others).

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America, leading to the positive phase in the Pacific–North American (PNA) teleconnection. Chen and Song (2000a,b) found that the early spring retreat of snow cover extent in Eurasia absorbs large amounts of heat from the atmosphere, leading to thermal anomalies in the troposphere, which ultimately induce anomalous circulation patterns over the mid-to-high latitudes of the Eurasian continent.

Several land factors such as soil moisture, soil temperature, and frozen soil have been considered to be the links with the long memories of land surface hydrological and thermal conditions and related feedback to climate (Shinoda 2001). Some studies have investigated the role of spring soil moisture (Matsuyama and Masuda 1998; Zuo and Zhang 2007, 2016; Zhao et al. 2007; Zhang and Zuo 2011) or frozen soil (Li et al. 2009) in the subsequent summer precipitation over East Asia and provide a theoretical basis for constructing the linkages. Chen and Song (2000a,b) suggested that the persistent soil moisture anomaly produced by early snow cover retreat from spring to summer appears to decrease the geopotential height field, which is not favorable for the development of blocking activity. Meanwhile, the soil moisture anomaly could also excite anomalous wave trains propagating from western Europe to East Asia through the geopotential height field, with opposite signs between northern and southern China. Using snow water equivalent (SWE) data, Wu et al. (2009a) examined the wave train response to reduced Eurasian spring SWE. Kripalani et al. (2002) and Wu et al. (2014) found that a west–east contrast pattern of the snow cover over the Eurasian continent is closely connected to summer precipitation anomalies over East Asia. Furthermore, Mu and Zhou (2012, 2015) demonstrated that the soil temperature and frozen soil anomalies associated with the extent of fresh snow in Eurasia might increase the northerly wind over the Baikal Lake (BL) region, leading to relatively colder conditions in northern East Asia, a strengthened subtropical upper-level jet stream, and a weaker Asian summer monsoon (Bamzai and Marx 2000; Dash et al. 2005).

The aforementioned studies are mainly based on snow cover variability itself. Spring is a transition season from winter to summer. Spring snow decrement (SSD) is closely related with snowmelt and latent heat sink and simultaneously reflects changes in soil moisture and the hydrological cycle, which influences the local thermal conditions and ultimately leads to climate anomalies both locally and remotely (Walsh et al. 1982; Groisman et al. 1994). Although these possible physical processes are identified, the detailed and quantitative climate impacts of Eurasian SSD on the East Asian climate have not been convincingly and deeply understood. Little attention has been paid to variations in the SSD and its relationship with the East Asian summer climate.

The purpose of the present study is to investigate the effect of Eurasian SSD on East Asian summer precipitation and to explore the associated dynamic and thermodynamic processes. The remainder of this paper is organized as follows. In section 2, we describe the data and methods used in the present study. Section 3 presents the relationship between Eurasian SSD and East Asian summer rainfall. Section 4 explores the associated dynamic and thermodynamic mechanisms. Section 5 discusses the role of El Niño and Arctic sea ice in the relationship between Eurasian SSD and East Asian summer rainfall, and conclusions are reported in section 6.

2. Data, methods, and model

The datasets used in this study include 1) the monthly mean atmospheric data from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis with a horizontal resolution of 2.5° × 2.5° available after 1948 (Kalnay et al. 1996); 2) the monthly mean sea ice concentration (SIC) and sea surface temperature (SST) from the Met Office Hadley Centre with a horizontal resolution of 1° × 1° available since 1870 (Rayner et al. 2003); 3) the monthly mean soil moisture from ERA-Interim with a horizontal resolution of 0.75° × 0.75° and four layers of 0–7, 7–28, 28–100, and 100–289 cm in depth for the period 1979–2013 (Balsamo et al. 2009), which can well represent the features of observed soil moisture over East Asia (Zuo and Zhang 2009; Liu et al. 2014); 4) the CPC gauge-based analysis of global daily precipitation with a horizontal resolution of 0.5° × 0.5° for the period 1979–2013 (Xie et al. 2007); and 5) the monthly mean snow water equivalent (SWE) data of GlobSnow version 2 (V2) from the Finnish Meteorological Institute for the period 1979–2013 (http://www.glbsnow.info/swe/), which are obtained using a combination of ground-based data and satellite microwave radiometer-based measurements and each monthly matrix having 721 × 721 grid (Solberg et al. 2009). We have converted the raw GlobSnow V2 SWE data to regular 1° × 1° grid for our analysis. For the majority of summer (June–September), measurable SWE is absent; thus, the data in summer are not credible and not utilized. In addition, SWE data are lacking south of 35°N; thus, the climate impact of SWE over the Tibetan Plateau is not discussed in our analysis.

In this study, the time period is taken from 1979 to 2013. The methods used in the present study include the empirical orthogonal function (EOF) as well as correlation and regression analyses. The Student’s t test is
used to check the statistical significance for the correlation and regression analyses.

The monthly snow decrement can be considered as the difference in SWE between two successive months. Positive and negative snow decrements represent increase and decrease SWE in succeeding month with respect to the previous one. Figure 1 shows the climatological annual cycle of SWE and snow decrement over the Eurasian continent (35°–76°N, 0°–150°E) during the period 1980–2013, with the absence of June–September SWE data. The monthly SWE rises considerably from October to the peak in February–March (approximately 61 mm) and then declines sharply in the following months to approximately 18 mm in May. Corresponding to the annual cycle of SWE, snow enhancement can be observed in cold seasons starting in October to the following February, followed by a hiatus in March; finally, the most significant snow reduction can be observed in April (−21 mm) and May (−24 mm). Therefore, the most severe snow reduction occurs mainly in the warming seasons in spring due to the gradually increasing temperature and decreasing snowfall. Therefore, in order to quantitatively depict the variations in snow decrement during the spring, in the present study, the Eurasian spring snow decrement is defined as the SWE difference between early spring (February–March) and late spring (May).

In this study, the NCAR Community Atmospheric Model, version 3.1 (CAM3.1), with a horizontal resolution of spectral T42 truncation (approximately 2.8° × 2.8°) and 26 vertical levels (up to 3.7 hPa) (Collins et al. 2006), was employed to better understand the dynamic and thermodynamic influences of Eurasian SSD on circulation anomalies over Eurasia, particularly East Asia and China. As shown in Collins et al. (2006) and Hack et al. (2006), this model is able to reproduce remarkably realistic atmospheric circulation and broad climatological characteristics of East Asian summer climate.

3. Results

a. Eurasian SSD and its relationship with East Asian summer precipitation

An EOF analysis was applied to the Eurasian SSD. The first three EOF modes of the SSD account for 22.2%, 14.3%, and 7.4% of the total variance, respectively. According to North et al. (1982), all these three modes are well separated from each other. Figure 2 shows the spatiotemporal features of the second EOF mode (EOF2) of Eurasian SSD. The spatial pattern of EOF2 is characterized by a west–east dipole pattern, with a negative center located in the region between eastern Europe and the West Siberia Plain (EEWSP) and a positive center in the vicinity of BL. Additionally, two other positive centers over eastern Europe and the North Siberia Plain are also identified (Fig. 2a). This dipole pattern somewhat resembles those of previous studies that are based on spring snow cover fraction (Yim et al. 2010; Wu et al. 2014). Wu and Kirtman (2007) also considered the West Siberia snow cover anomaly during spring to be a notable precursor for China seasonal precipitation. As shown by the principal component of EOF2 (PC2) (Fig. 2b), this pattern exhibits significant interannual and interdecadal variations. To quantify the Eurasian SSD variability and its association with the East Asian summer climate, we defined an SSD index (SSDI) as the difference in
area-averaged SSD between the domains with positive values (46°–58°N, 90°–130°E) and negative values (52°–66°N, 48°–80°E). The interannual and interdecadal variations of the SSDI are highly consistent with those of the PC2, and their correlation coefficient is as high as 0.91, indicating that the SSDI can provide a feasible representation of the spatiotemporal variations in Eurasian SSD during 1980–2013.

Corresponding to the variations in Eurasian SSD, the related East Asian summer precipitation anomaly is investigated by correlation analysis with the SSDI. As shown in Fig. 3a, the correlation coefficient pattern reveals a meridional quadrupole structure prevailing in the regions between northern and southern East Asia, with excessive precipitation over regions west of BL, northeastern China, and the Yellow River valley and deficient precipitation over Inner Mongolia and southern China. We checked the first three EOF modes of the East Asian summer precipitation. The first three modes account for 12.9%, 11.1%, and 8.0% of the total variance, respectively, and they are well separated from each other according to North et al. (1982). The meridional quadrupole pattern strongly resembles that of the spatial distribution of the third EOF mode (EOF3) of East Asian summer precipitation (Fig. 3b), and their spatial correlation coefficient is as high as 0.67. Furthermore, the variations in PC3 of the East Asian summer precipitation are similar to those of the SSDI during 1980–2013 (Fig. 3c), with the correlation coefficient being 0.49, which exceeds the 95% confidence level. The aforementioned indicates the significant relationship between the variations in Eurasian SSD and the East Asian summer precipitation.

However, there are also some discrepancies between the two precipitation patterns. The negative regions over Inner Mongolia in the EOF3 pattern appear to extend more northward toward BL and are more significant than those of the correlation pattern (Fig. 3).
These discrepancies illustrate that other factors may also have certain contributions, or their linkage may derive from the combined effects of SST (Zhang et al. 1999) and Arctic sea ice (Wu et al. 2009a). For instance, in 1998, the East Asian summer precipitation has negative anomalies larger than two standard deviations, whereas the SSD anomaly is insignificant (Fig. 3c), which may be related to Tibetan Plateau thermal forcing and east Pacific SST anomalies (Chinese National Climate Center 1998; Chen 2001; Lau and Weng 2001).

The above analysis reveals a statistical linkage between the Eurasian SSD and East Asian summer precipitation. In the next section, we will illustrate the physical mechanisms responsible for this linkage and associated causal relationship through exploring the possible dynamic and thermodynamic mechanisms associated with Eurasian SSD and East Asian summer precipitation, with a focus on surface thermal forcings.

b. Possible physical mechanisms

Previous studies suggest that Eurasian snowmelt may directly increase the soil moisture and affect soil temperature over some regions of Siberia during warming seasons, thereby leading to the reduction of surface air temperature and tropospheric diabatic heating (Cohen and Rind 1991; Matsuyama and Masuda 1998; Saito and Cohen 2003; Zuo et al. 2011). Figure 4 displays the regressed 100–289-cm soil moisture anomalies during late spring [April–June (AMJ)] and summer [June–August (JJA)] onto the SSDI. Figure 4a illustrates that the soil moisture anomalies feature an out-of-phase structure over the middle and high latitudes of the Eurasian continent during the late spring. In comparison, the western SSD anomalies are accompanied by stronger negative in situ moisture anomalies over EEWSP while the eastern SSD anomalies occur with positive soil moisture anomalies over regions around BL. The east–west contrast in soil moisture anomalies during late spring is consistent with the east–west contrast in the SSD anomaly pattern; thus, the soil moisture anomalies may be attributed to the effect of local spring SSD. The collocation of SSD and soil moisture anomalies supports the clear snow hydrological effect during the period 1980–2013, with excessive (less) SSD leading to more (less) snowmelt and thus wetter (drier) soil. Notably, the regions north of 70°N feature notable negative soil moisture anomalies, which is not in accordance with the excessive SSD during spring. In fact, these regions have the latest melt season due to their higher latitudes and the longest snowy seasons. Correspondingly, the soil moisture anomalies may not correspond well to the snow decrement during spring. In addition to the
mid-to-high latitudes, the majority of regions south of 35°N show coherent soil moisture anomalies. Because of the long memory of surface hydrologic conditions in Eurasia (Shinoda 2001), the Eurasian soil moisture structure persists to the subsequent summer. As shown in Fig. 4b, the pattern of the soil moisture anomalies in summer is quite similar to that in spring. In this study, we focus on the 100–289-cm layer soil moisture; other shallower soil levels display the coherent features but with relatively weaker signals (figures not shown), which may be attributed to the longer memory of soil moisture anomalies in deeper layer.

The soil moisture in summer (Fig. 4b) can affect the surface thermal balance. Zhang and Zuo (2011) examined the impact of soil moisture over East Asia on the surface heat balance and pointed out that a wetter (drier) soil corresponds to a lower (higher) surface air temperature. To analyze the effect of soil moisture on surface thermal condition in summer, we investigate the surface solar radiation flux, longwave radiation flux, sensible heat flux, latent heat flux, and net heat flux in regression onto the SSDI, and the results are shown in Fig. 5. The solar radiation at the surface is enhanced over EEWSP and weakened in the vicinity of BL (Fig. 5a); meanwhile, the longwave radiation varies oppositely (Fig. 5b). These effects may result from locally changed cloud cover in the two regions (Zampieri et al. 2009). The changed surface energy further results in more sensible heat but less latent heat in EEWSP, and less sensible heat and latent heat in the area around BL (Figs. 5c,d). A clear picture emerges in the net heat flux (Fig. 5e), which illustrates a similar anomalous pattern to that of the solar and sensible heat flux. The similarity of the anomalous net heat flux pattern to the distribution of the Eurasian SSD in spring as well as the soil moisture anomalies in spring and summer suggests a change of the local energy balance by the Eurasian SSD anomalies.

In Fig. 6 we show the atmospheric thermal features over Eurasian continent associated with the SSD. From Fig. 6a it can be seen that in summer over the dryer area in EEWSP significant higher surface air temperatures is observed, whereas over the wetter area around BL significant lower temperatures appears. These anomalous surface thermal forcings can warm or cool the lower troposphere. As shown in Fig. 6b, the near-surface thermal conditions associated with SSD substantially change the local atmospheric thickness of the lower–middle troposphere between 500 and 1000 hPa.

Fig. 4. (a) Late spring (AMJ) and (b) summer (JJA) 100–289-cm soil moisture (SM) anomalies (color shading; m$^3$ m$^{-3}$) regressed onto the SSDI. The rectangles here and in subsequent figures are as those in Fig. 2.
(Z500 – Z1000), with a notably thickened atmosphere over EEWSP and thinner atmosphere over regions west of BL. The west–east dipole pattern in atmospheric thickness is quite similar to those in surface temperature and soil moisture. The meridional heat flux anomalies over East Asia shown in Fig. 5e form the north–south dipole structure in temperature anomalies over East Asia. This dipole further strengthens the meridional temperature gradient and atmospheric baroclinicity over the midlatitude of East Asia at 700 hPa and simultaneously decreases them to the north and south (Fig. 6c). These results reflect the thermal processes associated with anomalous Eurasian SSD, and illustrate the possible thermodynamic influences on atmospheric circulation over East Asia in summer.

According to the thermal wind adjustment theorem, the intensified atmospheric baroclinicity can accelerate the 200-hPa westerly winds over the midlatitudes of the Eurasian continent, leading to a strengthened subtropical jet stream over northern China and easterly wind anomalies north and south of the subtropical jet stream (Fig. 7a). The strengthened subtropical jet tends to act as a connection between near-surface thermal forcings and the development of meridional circulation anomalies, as suggested by Mu and Zhou (2012).

Figure 7b illustrates that at the 500-hPa geopotential heights, an anomalous cyclonic pattern and anomalous anticyclonic pattern emerge to the north and south, respectively, of the exit region of the East Asian subtropical jet stream. In addition to this dipole pattern, an evident midlatitude Eurasian wave train also prevails over the regions from the West Siberia Plain to the northwestern Pacific, with a negative center located over Mongolia and positive centers over EEWSP and the northwestern Pacific, respectively. This Eurasian wave train can be interpreted as a manifestation of a single stationary Rossby wave train propagation. Meanwhile, coexisting with the midlatitude wave train, in the higher latitudes a relatively weaker polar wave train with weakened (strengthened) heights emerges over regions around the western (eastern) Siberian coast. These results imply that the anomalous SSD-induced upper-level westerly anomalies can provide favorable dynamic conditions for the development of a north–south dipole pattern and Eurasian wave train pattern over the lower-middle and higher latitudes of East Asia, which tend to act as the atmospheric bridge linking the surface thermal forcings and East Asian summer climate.

In addition to the influences of the wave trains discussed above, the SSD and/or soil moisture anomalies
may also directly trigger the wave trains during late spring (AMJ) and summer. The origins and development of the Eurasian wave trains are demonstrated in Fig. 8, which presents the 500-hPa streamfunction anomalies and associated wave activity flux (WAF; Takaya and Nakamura 2001) during late spring and summer. Over EEWSP, significant positive streamfunction anomalies associated with reduced SSD and strengthened thermal heating occur in situ during late spring. Meanwhile, there are relatively weak WAFs spreading from extratropical North Atlantic northeastward to eastern Europe along the subpolar belt. This wave train is then markedly reinforced over eastern Europe, propagating northeastward to the West Siberia Plain, and then splitting into two branches, with one turning to the higher latitudes and another southeastward to East Asia (Fig. 8a). This reinforced propagation pattern can be attributed to the thermal forcing associated with SSD anomalies. In summer, similar and significant midlatitude WAFs emanate from eastern Europe and propagate directly eastward to BL and farther to East Asia, exhibiting predominantly zonally oriented. The horizontal propagation of quasi-stationary wave train in summer (Fig. 8b) resembles
the Silk Road and polar wave trains as suggested by Orsolini et al. (2015) that have substantial impacts on precipitation over northeast China. Compared with those during late spring, in summer the upper-stream wave train across the extratropical North Atlantic is much weaker and the wave propagation is more zonally oriented. This implies that the Eurasian wave trains are essentially triggered by SSD anomaly over EEWSP, and the persistence of the wave propagation is mainly related to the consecutive surface thermal forcings.

Some recent studies have emphasized the interference between the forced and climatological wave trains (Smith et al. 2011; Smith and Kushner 2012; Peings and Magnusdottir 2014; Orsolini et al. 2015). These studies suggest the importance of both forced and climatological wave trains in affecting the atmospheric circulation. For instance, the atmospheric circulation is stronger when the forced wave train is in phase with the climatological one, and vice versa. Figure 9 plots the forced anomalous wave train regressed on the SSDI along with the climatological wave train, which are represented by 500-hPa geopotential heights. The Fourier decomposition is used to decompose the stationary waves from wave 1 to wave 4 according to zonal wavenumber.

In the original field (Fig. 9a), the climatological wave trains in summer exhibit the coexistence of the polar and midlatitude wave trains. The forced and climatological wave trains are in phase over EEWSP and regions south of BL, and out of phase over regions north of BL and around the Okhotsk Sea, with a spatial correlation coefficient of 0.17. Over EEWSP the spatial correlation coefficient is as high as 0.90, which denotes a constructive linear interference between the forced and climatological wave train. Concerning wave 1 and wave 3 (Figs. 9b,d), the forced wave train is out of phase with the climatological one with the spatial correlation coefficients being −0.53 and −0.44 for waves 1 and 3, respectively. These destructive interferences indicate that the atmospheric circulation modified by waves 1 and 3 is opposite to the climatological wave train. As for waves 2 and 4 (Figs. 9c,e), their spatial correlation coefficients with the climatological wave train are 0.36 and 0.41, respectively, suggesting that the anomalous forced waves should be linearly interfered with and enhance the climatological wave train. Thus, the constructive linear interferences exhibited in waves 2 and 4 would reinforce the climatological wave train and then prompt the influence on East Asian summer climate.
These anomalous circulation patterns provide favorable conditions for summer precipitation over East Asia, particularly over China, with excessive precipitation over regions west of BL, northeastern China, and the Yellow River valley, and deficient precipitation over Inner Mongolia and southern China (Fig. 3a). These results indicate that the Eurasian SSD-induced diabatic heating forcings may make prominent contributions to East Asian summer precipitation through the formation and development of the Eurasian wave trains. However, these results depend largely on observation analyses and cannot guarantee a causal relationship. Considering that the numerical model simulations are essential to confirming the diagnostic results, the simulations are addressed in the next section.

c. Model simulations

To elucidate the influence of anomalous Eurasian SSD on East Asian summer climate, we performed a set of control experiments with CAM3.1 to derive different initial fields. The model is spun up for 20 yr and then integrates forward for 50 yr. The data for 1 December of the last 50 yr are used as initial fields, and the 50 initial conditions are used for ensemble simulation. Then, the control and sensitivity experiments, each with 50 members, are integrated from the beginning of December to the end of August in the following year. In the sensitivity experiments, we refer to Barnett et al. (1989) for the experimental design. To quantitatively mimic the observed Eurasian SSD anomalies (as shown in Fig. 2), we reduce the snowfall rate by 50% over EEWSP and simultaneously increase the snowfall rate by 50% over regions around eastern Europe, the North Siberia Plain, and BL. Such modification is applied to the entire period of the sensitivity simulations. Finally, the simulation results are obtained as the ensemble difference between the sensitivity and the control experiments. To increase the robustness and reliability of the numerical experiments, a 50-member ensemble mean is conducted to reduce the uncertainties arising from differing initial conditions. Corresponding to the change in snowfall rate, other hydrological and energy variables in CAM3.1 simulations change accordingly. In both late spring and summer, the deficient snowfall is accompanied by negative anomalies of 100–289-cm layer soil moisture and lower temperature over EEWSP, while more snowfall is accompanied by positive soil moisture anomalies and lower temperature over regions around BL. These anomalous patterns are similar to the Eurasian SWE and SSD anomalies over Eurasia (Fig. 10).

To substantiate the conjecture of our designed experimental schemes, we primarily examined the model-simulated response of spring SWE and SSD to the anomalous snowfall rate. As shown in Fig. 10a, two regions of large and significant negative anomalies are induced in the surface SWE field during spring, one located over EEWSP and the other over the Tibetan Plateau, with the maximum value occurring in the latter.
region and exceeding −8 mm. This reflects the in-phase variation in SWE over the two regions and offsets the absence of observed lower-latitude SWE. Similarly, three positive SWE anomalous regions are also triggered, with centers emerging over eastern Europe, the North Siberia Plain, and BL. The maximum positive SWE center occurs in regions west of BL, with the center value exceeding 8 mm. Corresponding to the simulated anomalous SWE, the SSD reveals a similar pattern, namely more SSD over eastern Europe and BL and less SSD over EEWSP and the northern Tibetan Plateau (Fig. 10b). However, there are some differences between SWE and SSD. One is that the largest negative SSD center occurs over EEWSP at approximately 60°N, which differs from that of the SWE at approximately 70°N. Another is that the negative Tibetan Plateau SSD center is much weaker than that of the SWE and even a positive center can appear, implying weak snowmelt and snow enhancement over the northern and southern Tibetan Plateau, respectively, possibly due to the lower temperature over the plateau. Moreover, an insignificant negative SSD anomaly is simulated over the North Siberia Plain, perhaps due to the persistent snowfall, even during early summer. These simulations largely reproduce the observed anomalous SSD patterns (Fig. 2a), indicating that our numerical schemes of the snowfall rate changes are representative of the SSD anomalies in the CAM3.1.

To verify whether the anomalous SSD can induce the Eurasian wave train, we simulated the physical processes that determine the circulation in response to snowfall rate changes. Figure 11 presents the simulated composite differences between sensitivity and control experiments in those fields parallel to reanalysis (Figs. 6–8). The surface air temperature field (Fig. 11a) shows a positive difference over eastern Europe and a negative difference over BL, reproducing the west–east dipole pattern in reanalysis. In the atmospheric thickness field (Fig. 11b), two enhanced anomalous thickness centers appear over eastern Europe and central China, whereas a decreased thickness center emerges over regions north of BL. Meanwhile, in the 700-hPa meridional temperature gradient field (Fig. 11c), an anomalous baroclinic center appears to the south of BL. These patterns respond directly to the thermal forcing of Eurasian SSD. Then, a strengthened subtropical jet stream (Fig. 11d), an Eurasian midlatitude wave train, and a north–south dipole structure over East Asia with...
decreased height over BL and an enhanced height over southern China (Fig. 11e) are subsequently simulated. Figure 11f shows weak WAFs propagating from extratropical North Atlantic eastward to East Asia, which is markedly reinforced in eastern Europe, indicating the active role of SSD anomaly over EEWSP in triggering a strong Rossby wave train.

The linear interference between the model-forced and climatological wave trains is further examined in Fig. 12. In the total wave (Fig. 12a), the forced and climatological waves are in phase over EEWSP with the spatial correlation coefficient being 0.83 and out of phase over regions north of BL. Concerning the decomposed waves, the forced wave is out of phase with the climatological wave for waves 1 and 3 with spatial correlation coefficients being −0.23 and −0.36, respectively, and in phase for waves 2 and 4 with spatial correlation coefficients being 0.23 and 0.39, respectively. These responses further confirm that the linear interference between the forced and climatological waves in waves 2 and 4 intensifies the influence on East Asian summer climate.

In general, these simulations reproduce the thermodynamic and dynamic responses of atmospheric circulation to anomalous Eurasian SSD as identified in the analysis using observation data, and demonstrate the previous argument that the anomalous Eurasian SSD can excite a Eurasian wave train propagating from eastern Europe eastward to East Asia. However, there are still some discrepancies in the simulated geopotential height response for the polar wave train compared to that in reanalysis. For example, the total wave response (Fig. 12a) is farther north than in the reanalysis (Fig. 9a). Examining Fig. 12d, it seems that the model has a stronger forced response at high latitudes than the reanalysis (Fig. 9d). In Fig. 12e, the response is slightly shifted north compared to reanalysis (Fig. 9e). These discrepancies may be due to the weak SSD forcing over high latitudes in CAM3.1 or contributions from other factors such as North Atlantic SST (Wu et al. 2011) and Arctic sea ice (Wu et al. 2013).

The associated summer precipitation anomalies are also shown in Fig. 13. Positive precipitation anomalies are identified over regions from BL eastward to northeastern China and the lower Yellow River valley, and negative anomalies over western Inner Mongolia and most regions of southern China. When compared with the observed anomalous precipitation patterns in Figs. 3a and 3b, opposite signs appear in
regions from northern China to the middle Yellow River valley. These discrepancies may arise from the deficiency of CAM3.1 in simulating variations in East Asian summer precipitation, as described in Hack et al. (2006) and Wei et al. (2011). Nevertheless, the numerical simulations demonstrate the significant role of Eurasian SSD in influencing the East Asian summer precipitation through modulating atmospheric circulation pattern.

4. Discussion of other factors' effects

Although the close connection between Eurasian SSD and East Asian summer climate has been examined with both observations and numerical model simulations in the above sections, other atmospheric factors and lower boundary forcings, such as ENSO (Zhang et al. 1996, 1999; Ferranti and Molteni 1999; Shaman and Tziperman 2005) and Arctic sea ice (Wu et al. 2009b,c; Zhang and Wu 2011; Bader et al. 2011; Liu et al. 2012), can act as important drivers of variations in both Eurasian snow cover and East Asian summer climate. In this section, we will employ partial correlation analyses (Spiegel 1988) to examine if the relationship between Eurasian SSD and East Asian summer climate revealed in our present study is affected by ENSO and Arctic sea ice.

Figure 14 shows the partial correlation coefficient patterns between the SSDI and East Asian summer precipitation after the linear parts related to Niño-3.4 index and SIC index are removed, respectively. The Niño-3.4 index is defined as the average of eastern equatorial Pacific sea surface temperature anomalies in the area 5°S–5°N, 120°–170°W. The SIC index refers to the regional averaged SIC north of 60°N. The partial correlation patterns feature similar structures to those of the EOF3 pattern of East Asian summer precipitation (Fig. 3b) and the correlation coefficient pattern (Fig. 3a), with anomalous meridional quadrupole precipitation centers extending from northern to southern East Asia. To further examine their resemblance, we perform a spatial correlation analysis to confirm the resemblances between the partial correlation coefficient patterns and EOF3 pattern of East Asian summer precipitation. The spatial correlation coefficients between the EOF3 pattern of East Asian summer precipitation and the partial correlation coefficient patterns with ENSO and SIC removed are 0.65 and 0.68, respectively, which are close.

Figure 11. CAM3.1 simulations of summer (a) surface air temperature (color shading; °C), (b) atmospheric thickness (Z500 − Z1000; color shading; gpm), (c) 700-hPa meridional temperature gradient (color shading; °C grid−1), (d) 200-hPa zonal wind (color shading; m s−1), (e) 500-hPa geopotential height (color shading; gpm) anomalies, and (f) 500-hPa streamfunction (color shading; m² s−1) and associated wave activity flux (vectors; m² s−1) anomalies.

FIG. 11. CAM3.1 simulations of summer (a) surface air temperature (color shading; °C), (b) atmospheric thickness (Z500 − Z1000; color shading; gpm), (c) 700-hPa meridional temperature gradient (color shading; °C grid−1), (d) 200-hPa zonal wind (color shading; m s−1), (e) 500-hPa geopotential height (color shading; gpm) anomalies, and (f) 500-hPa streamfunction (color shading; m² s−1) and associated wave activity flux (vectors; m² s−1) anomalies.
to 0.67 between the correlation pattern (Fig. 3a) and the EOF3 pattern of East Asian summer precipitation (Fig. 3b). Therefore, the relationship between Eurasian SSD and East Asian summer precipitation reveal in our present study should be independent of ENSO and Arctic sea ice concentration.

5. Conclusions

In the present study, we investigated the relationship between Eurasian spring snow decrement (SSD) and East Asian summer precipitation and the related thermodynamic and dynamic mechanisms using both observational data and the CAM3.1 model. The results show that the second EOF mode of Eurasian SSD exhibits a west–east dipole pattern, with a negative center located over eastern Europe and the West Siberia Plain (EEWSP) and a positive center over the area around Baikal Lake (BL). This anomalous SSD pattern is significantly associated with the third EOF mode of East Asian summer rainfall through triggering an anomalous midlatitude Eurasian wave train. The reduced SSD over EEWSP tends to decrease the local soil moisture from spring to the following summer, thereby increasing the surface heat flux and near-surface temperatures. Similarly, the increase in SSD over BL is accompanied by anomalously low near-surface temperatures. Changes in near-surface temperatures further intensify the meridional temperature gradient and lower-level baroclinicity, leading to the acceleration of the upper-level subtropical westerly jet stream. At 500-hPa geopotential heights, an anomalous cyclone and anomalous anticyclone emerge to the north and south, respectively, of the exit region of East Asian subtropical jet stream. Meanwhile, the changed surface thermal conditions enhance the local 1000–500-hPa thickness over EEWSP while decreasing it over BL. These factors both create favorable physical conditions for the maintenance and enhancement of the anomalous midlatitude Eurasian wave train prevailing over the regions from eastern Europe eastward to the northwestern

![Fig. 12. As in Fig. 9, but for CAM3.1 simulations.](image1)

![Fig. 13. CAM3.1 simulations of summer precipitation anomalies (color shading; mm).](image2)
Pacific. We further explored the origin of the Eurasian wave train and found that there are zonally oriented WAFs spreading from eastern Europe eastward to East Asia. This finding demonstrates the role of anomalous SSD in triggering the Eurasian wave train. These circulation patterns ultimately significantly influence the precipitation over East Asia, especially over China, with excessive precipitation over regions west of BL, northeastern China, and the Yellow River valley and deficient precipitation over Inner Mongolia and southern China. Therefore, our study confirms the significant role of Eurasian SSD in influencing East Asian summer precipitation.

Our model results demonstrate that the CAM3.1 can reproduce the positive summer precipitation anomalies over most of northern East Asia and negative anomalies over southern China. Corresponding to the imposed anomalous Eurasian SSD forcings in CAM3.1, positive surface air temperature and atmospheric thickness responses over EEWSP and negative responses over BL are primarily simulated. In the simulation, the subtropical jet stream over East Asia is strengthened, the Eurasian midlatitude wave train is formed, and reinforced WAFs propagate from eastern Europe to East Asia. These simulated dynamic and thermodynamic processes result in according precipitation anomalies to occur over East Asia.

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REFERENCES


